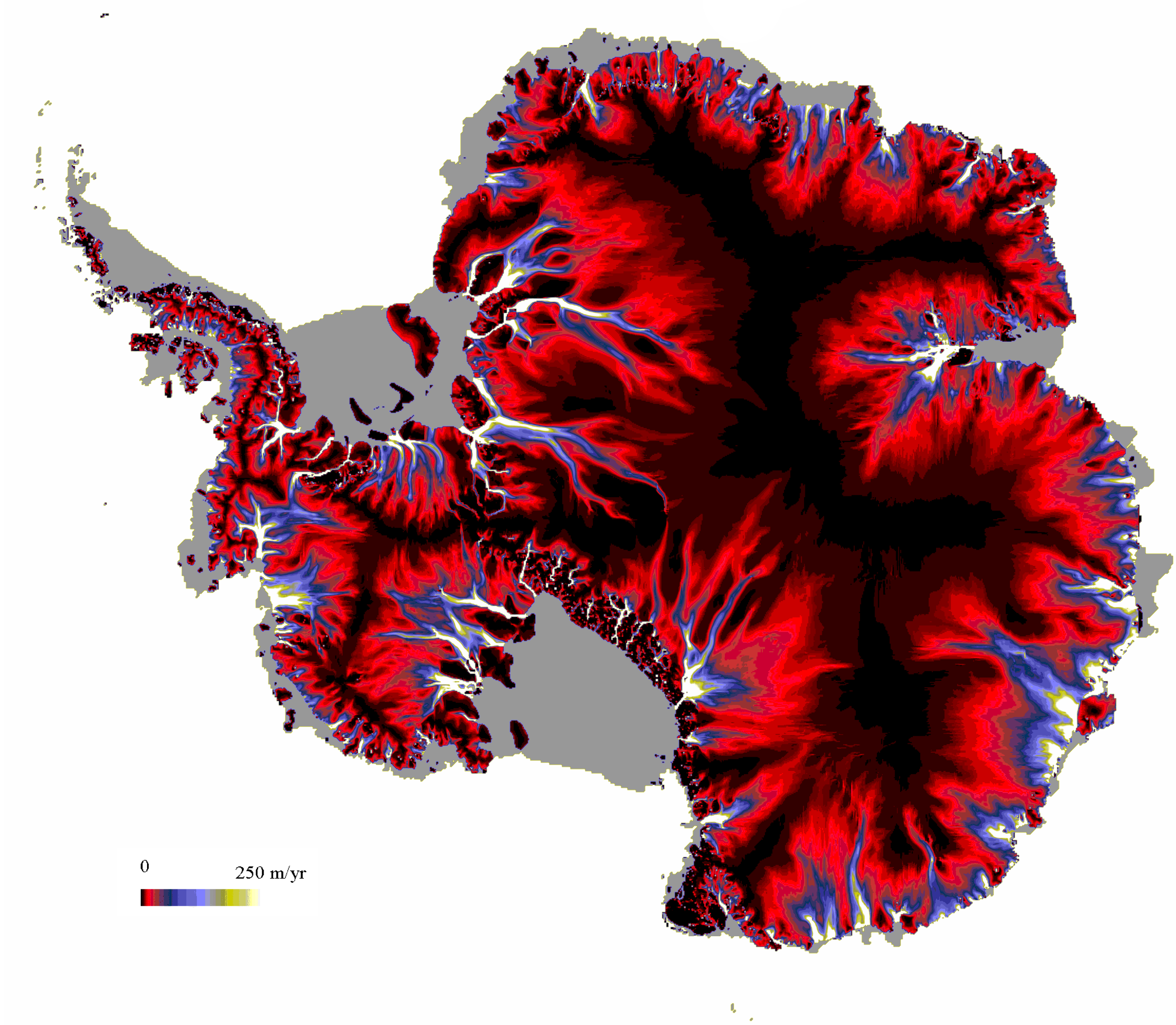
LITERATURE REVIEW

WEST ANTARCTIC ICE STREAMS – A REVIEW OF THE PROPOSED MECHANISMS OF ICE STREAMING



GCAS 2008/09 | Martina Armstrong

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1.1 Introduction

The Antarctic Ice Sheet, which covers most of the continent's land surface, is divided into two separate entities by the Transantarctic Mountains. These are regarded as the West and East Antarctic Ice Sheets (Figure 1). The West Antarctic Ice Sheet (WAIS) contains 3.8 million km^3 of ice and, unlike the East Antarctic Ice Sheet (EAIS), is a marine ice sheet which implies that the grounding-line is below sea-level (Figure 2). Weertman (1974) originally proposed that a marine ice sheet, such as the WAIS, is inherently unstable. However, this analysis was based on a simple model of a marine ice sheet that did not include fast-flowing, wet-based ice streams, which are now known to dominate the grounded ice sheet (Bentley, 1998). It has been estimated that the ice streams on the WAIS move ~10-100 times faster than the adjacent non-streaming ice sheet (Bindschaler and Scambos, 1991; Whillans and van der Veen, 1993). Swithinbank (1954) defines an ice stream as part of an inland ice sheet in which the ice flows more rapidly than, and not necessarily in the same direction as, the surrounding ice. This definition indicates two main points: (1) ice streams are surrounded by ice as, if it were surrounded by rock, it would be considered an outlet glacier, and (2) it is part of the inland ice sheet, therefore it is not floating.

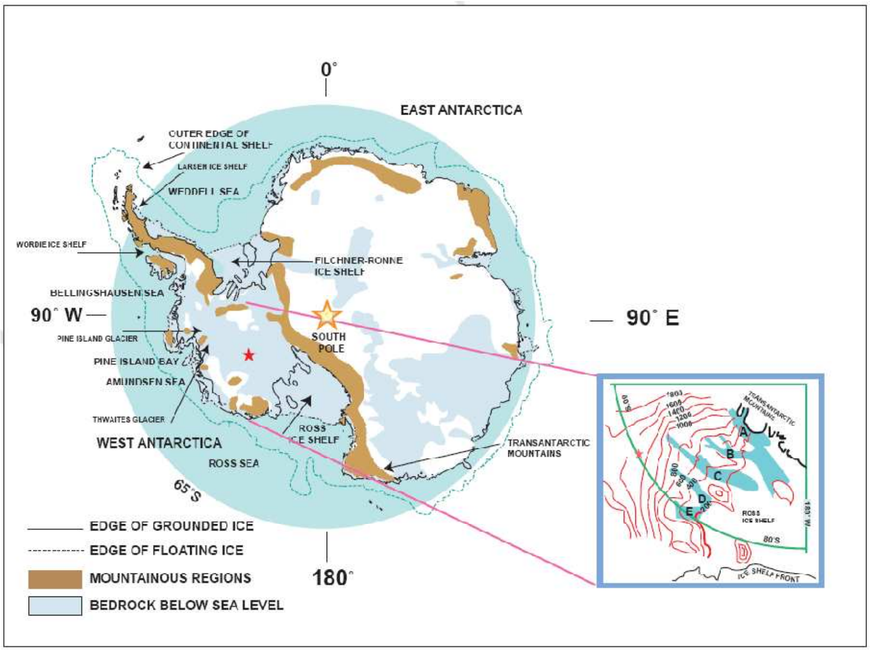

Figure 1: The Antarctic Ice Sheet. WAIS, located to the left of the Transantarctic Mountains, is largely grounded below sea level. Ice streams (A to E) draining into the Ross Ice Shelf are shown at lower right; ice elevation is in metres above sea level. The red star in both maps, Byrd Station, is provided for orientation. WAIS as defined here includes the Ross and Filchner-Ronne ice shelves but not the Antarctic Peninsula (top left). The locations of mountainous regions, deep basins, margins of floating ice, and grounding lines at ice shelves are indicated approximately. The Transantarctic Mountains are divided in places by deep basins not shown here so that ice from both East and West Antarctica drains into the large ice shelves. *Source:* Oppenheimer (1998).

The stability of the WAIS, and its associated ice streams, has come to the forefront of glaciological research in recent years due to the well-publicised climate change debate (Oppenheimer, 1998). The WAIS has the potential to contribute to rapid sea-level rise, affecting a vast number of livelihoods, which is why there is a need to fully understand the mechanisms behind ice sheet collapse and use this to quantify the risk of such an event occurring. Understanding the dynamics of the West Antarctic ice sheet (WAIS) requires comprehension of the processes and interactions that control the development and movement of ice streams (Blankenship et al., 1986; Alley et al., 1986). In particular, behaviour in the basal zone of ice sheets has been attributed to the development and movement of ice streams in numerous research papers (e.g. Blankenship et al., 1986; Engelhardt et al., 1990; Bentley et al., 1998; Kamb 2001). This review seeks to outline the proposed mechanisms for ice streaming that operate in the basal zone of the WAIS – Ross Ice Shelf System (RISS).

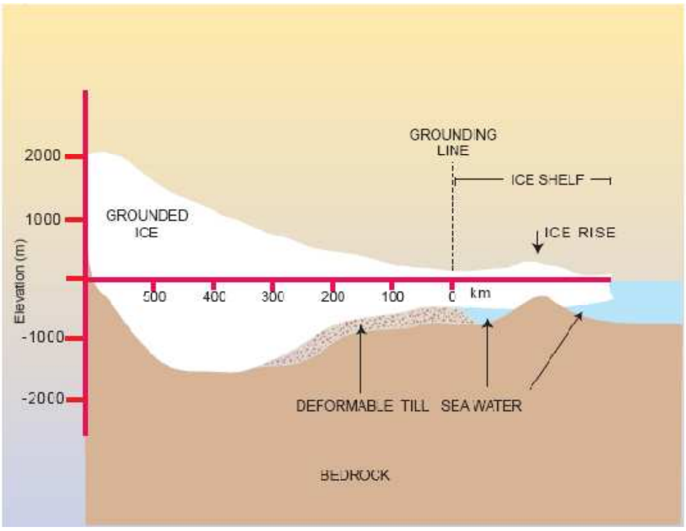

Figure 2: Cross-section of an ice stream and ice shelf of a marine ice sheet, indicating location of grounding line, bedrock rise on the ocean floor, and possible extent of deformable till. The thickness of the till layer, actually a few metres, is exaggerated for clarity. *Source:* Oppenheimer (1998).

2.1 The Ice Streams of the West Antarctic Ice Sheet – Ross Ice Shelf System

Ice discharge from the WAIS is dominated by ice streams that feed into the two largest ice shelves in existence: the Ross and Filchner-Ronne. About half of the ice flows into the Ross Ice Shelf (Figure 1). There are five ice streams in this drainage basin, with each measuring 30–80 km wide and 300–500km long (Oppenheimer, 1998). From south to north along the west side of the WAIS, the ice streams are named A though E, however, new names have been applied to them in recent years.

Ice stream A (Mercer Ice Stream), which is located at the foot of the Transantarctic Mountains, drains ice from the high East Antarctic plateau via Reedy Glacier and the Shimizu Ice Stream. Deep channels are aligned with the two outlet glaciers, with the remaining bedrock being comparatively rough. Ice stream B (Whillans Ice Stream) converges with A into a stream more than 100 km wide near the grounding line. The lower half of ice stream C (Kamb Ice Stream) ceased flowing rapidly about 140 years ago (Rose, 1979; Retzlaf and Bentley, 1993). However the upper section is still moving at ice-stream-like speeds (Fahnestock and Bamber, 2001). Ice stream D (Bindschadler Ice Stream) drains from the area of Byrd Station and near the Ross Ice Shelf it joins with ice stream E (MacAyeal Ice Stream). The catchment area of ice stream E extends north to the ice divide connecting the coastal ranges of Marie Byrd Land. The trunks of ice streams A, B, D, and E are characterised by low surface slopes, high flow speeds, and intense marginal shear (Fahnestock and Bamber, 2001). Ice stream C appears to have had similar characteristics when it was flowing rapidly.

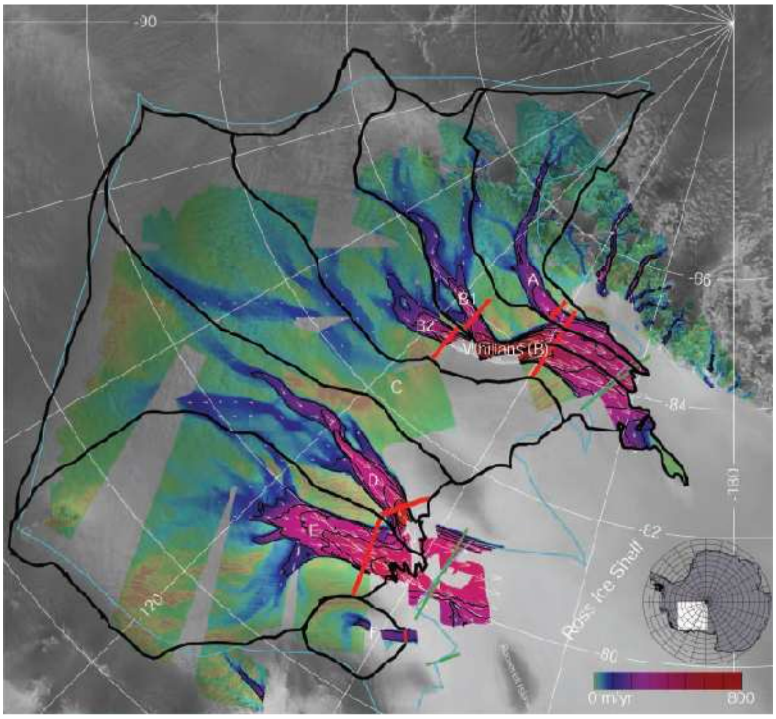

Figure 3: Ice-flow velocity (colours) over radar imagery from the RADARSAT Antarctic Mapping Project Mosaic. Flow velocity at 100 m/year intervals is contoured with thin black lines. White vectors show subsampled velocity vectors in fast-moving areas. Catchment boundaries for individual ice streams are plotted with thick black lines. Flux gates used in discharge calculations are shown with red lines. *Source:* Joughin and Tulaczyk (2002).

The initial overview of these ice streams came from aircraft-based radio echo-sounding, aerial photography, and early Landsat imagery (Rose, 1979). The WAIS itself lies in a deep, sediment-laden marine basin warmed from beneath by a relatively high geothermal heat flux (Bindschadler et al., 2001). These factors combine to create the potential for a well-lubricated, water-saturated bed on which ice streams can maintain fast motion through the generation of sufficient quantities of water by basal friction and shear heating. This is despite the fact that the driving stresses for these ice streams are far less than those acting on a typical outlet glacier (Joughin et al., 2002). Joughin and Tulaczyk (2002) calculated both the ice-flow velocities (Figure 3) and the associated discharge and accumulation fluxes (Table 1) for the individual Ross Ice Streams. Strong evidence for ice sheet growth (+26.8 gigatons per year) was found; this was in contrast to earlier estimates indicating a mass deficit (-20.9 gigatons per year) (Shabtaie and Bentley, 1987). The Siple Coast ice streams in particular have exhibited very large flow fluctuations (Fahnestock et al., 2000), including the aforementioned stagnation in flow of Kamb Ice Stream around 150 years ago (Retzlaff and Bentley, 1993), and the current slowdown of Whillans Ice Stream (Joughin et al., 2002).

Table 1: Discharge and accumulation fluxes for the Ross Ice Streams. The summed total excludes the first three entries because the combined flux through these gates is included by the wider gate farther downstream (Figure 3). *Source:* Joughin and Tulaczyk (2002).

Gate	Accumulation above gate (10^{12} kg year^{-1})	Discharge flux (10^{12} kg year^{-1})	Difference (10^{12} kg year^{-1})	Area above gate (km^2)	Average accumulation rate (kg m^{-2} year^{-1})
Ice Stream A	10.1 (1.5)	8.0 (0.3)	2.10 (1.5)	72,400	139
Whillans B1	8.0 (1.2)	12.7 (0.4)	−4.70 (1.3)	52,400	153
Whillans B2	12.9 (1.9)	10.3 (0.4)	2.60 (2.0)	87,900	146
Ice Stream A and Whillans Ice Stream	32.8 (4.9)	30.3 (0.7)	2.50 (5.0)	235,200	139
Ice Stream C	20.5 (3.1)	0.5 (0.2)*	20.1 (3.1)	153,400	134
Ice Stream D	19.0 (2.9)	15.3 (0.4)	3.75 (2.9)	140,300	136
Ice Stream E	23.5 (3.5)	24.4 (0.7)	−0.85 (3.6)	175,200	134
Ice Stream F	2.8 (0.4)	1.5 (0.1)	1.35 (0.4)	16,800	171
Total AW+C+D+E+F	98.8 (14.8)	72.0 (1.1)	26.8 (14.9)	721,000	137
	102.9 (15.4)		30.8 (15.4)		142
	94.7 (14.2)		22.7 (14.2)		131

3.1 Mechanisms of Ice Streaming

Studies on the ice streams in the Siple Coast have provided a vast range of knowledge on the behaviour of the basal zone of ice streams. As stated earlier, without the possession of knowledge about these mechanisms the quantification of the risk of ice sheet collapse cannot be made. A variety of methods are used to determine the mechanism of ice flow. The subglacial properties of the ice streams that drain into the Ross Ice Shelf have been studied using boreholes and seismic methods (Oppenheimer, 1998).

There has been much debate over the cause of ice streaming. Earlier suggestions included ice superplasticity (Hughes, 1977) where it was believed that the viscoplastic instability of ice and subglacial topography were responsible for the formation of ice streams near ice sheet margins grounded below sea level. Viscoplastic instability is inherent in anisotropic polycrystalline solids, such as glacial ice, therefore this theory is valid. The influence of the temperature of ice on its viscosity reinforces the tendency for fast ice flow to be associated with bedrock troughs. Ice occupying troughs is likely to be relatively warm both because of the insulating effect of thicker ice and because fast-flowing ice generates heat by internal dissipation (Payne, 1999). Trough ice will therefore be warmer, have a reduced viscosity and flow faster than adjacent non-trough ice. Ice streams A-E only have loose topographic control, therefore are not affected by this. Air-borne radar studies (Shabtaie and Bentley 1987) indicate that although these ice streams follow bedrock troughs, the relationship with the underlying topography is complex with many troughs being unoccupied and ice streams lying asymmetrically within their troughs. The apparent lack of a direct topographic control has encouraged a search for other controlling mechanisms. Continued research on ice streaming provided two more theories: basal sliding due to basal melting and lubrication of the bed by soft sediment.

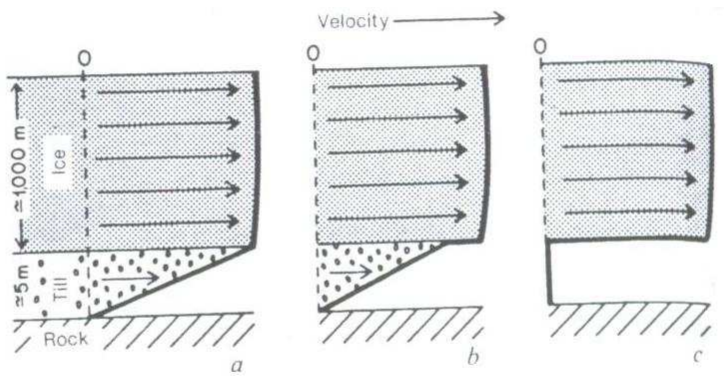

Figure 4: Possible models for an ice stream bed. *a*, Till deformation only; *b*, Till deformation plus basal sliding; *c*, Basal sliding only. *Source:* Alley et al. (1986).

3.2 Till Deformation vs. Basal Sliding

An early paper by Alley et al. (1986) proposed that deformation within the till deposits underlying the ice sheet can be considered as the primary mechanism by which ice streams move (Figure 4a). This has since been disputed by various authors; however, it is still important to look at the evidence that Alley et al. (1986) used to support such a statement. Their research was undertaken on upstream B where the evidence consisted of three factors: (1) till porosity, (2) force balance, and (3) water balance. Boulton and Paul (1976), in a separate geotechnical study, determined that lodged basal till has a porosity of $\lesssim$ 30% and that the deformation of till causes dilation and porosity of ~ 40%. Seismic evidence from Alley et al. (1986) found that the basal till located at upstream B had a porosity of ~ 40%. These findings have also been supported by research undertaken on upstream B by Engelhardt et al. (1990). This value was too high for lodged till, therefore it was considered consistent with active deformation. It must be noted that porosities in ice stream D are on average about 8% higher than in B or C, indicating that the till at ice stream D has a lower strength and, therefore, contribute even less basal drag to the driving stress of ice stream flow (Kamb, 2001).

Active till deformation at upstream B was also evident when the driving stress (τ_d) for ice flow at upstream B was calculated. The initial driving stress (τ_d), or force balance, behind ice streaming can be given by:

$$\tau_d = \rho_i \, gh \, \sin \alpha$$

where:
ρ_i is the density of ice,
g is the acceleration of gravity,
h is the ice thickness, and
α is the surface slope.

Source: Bentley (1987).

The driving stress at this location, where ρ = 920 kg m^{-3}, g = 9.8 m s^{-2}, $h \simeq$1,050 m, and $\sin \alpha$ = 0.002, was calculated at ~ 19 kPa. However, it was recognised by Alley et al. (1986) that the driving stress is resisted by gradients in stretching stress and by side drag as well as by the basal drag. A force balance equation for the ice stream indicated that basal drag is the major resistive force. Therefore, it was concluded that 95% of the driving stress was balanced by basal drag at upstream B, making the basal drag 18 kPa. This was more than twice the best estimated value of the strength of dilated till, so they expected till deformation to be occurring at this location.

Since this time, Whillians and van der Veen (1997) have found, through aerial photogrammetry, that the bed under Ice Stream B is very weak and unable to provide much resistance; therefore mechanical control on this ice stream originates almost entirely from the lateral margins. This is also supported by earlier work by Blankenship et al. (1986) where seismic evidence suggested that, at least locally, the bed under Ice Stream B consists of a weak water-saturated layer of till, unable to resist ice stream flow. In saying this, however, the strength of dilatant till can increase when the void ratio is decreased by removing water from the till, which can result from basal freezing or basal drainage (Tulaczyk et al., 2000). The graph in Figure 5 shows the empirical relationship between the strength of till at upstream B and the void ratio.

Based on a comparison to findings at Breidamerkurjökull by Boulton (1979), Alley et al. (1986) also assume that the entire thickness of till was deforming at upstream B. They expected that the strain rate equalled the minimum strain rate required to cause stress concentrations large enough to mobilise lodged till. However, Engelhardt and Kamb (1998), using a tethered stake instrument in a borehole at upstream B, found that the shear deformation of till is within 3 cm of the ice-bed interface. This indicates that Alley et al. (1986) have over-estimated the strain rates active at upstream B. In particular, it may relate to their proven error (e.g. Blankenship et al., 1986; Whillians and van der Veen, 1997) in estimating the resistive force of basal drag in the driving stress calculations.

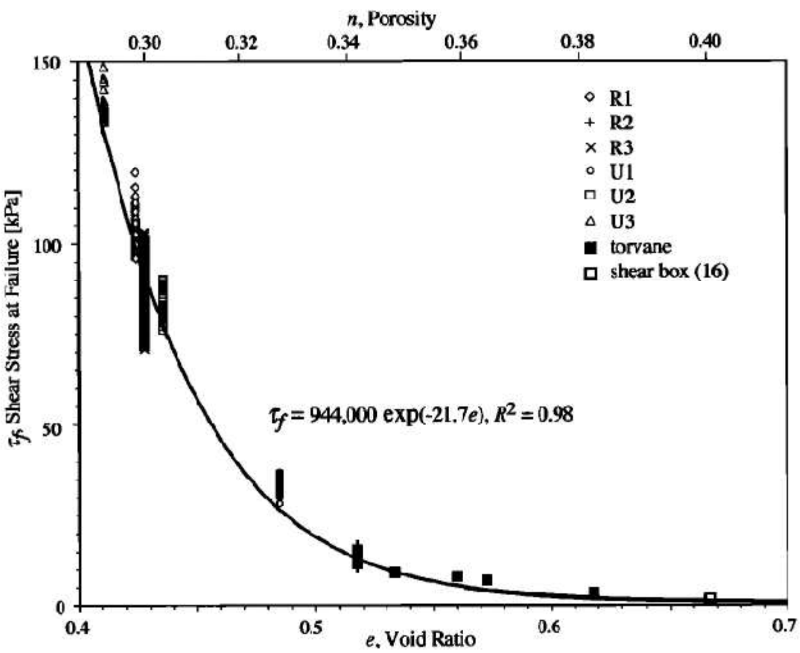

Figure 5: Empirical relationship between the strength of the UpB till τ_f and void ratio e. The thick solid line and the equation give the best fit relationship derived from results of six undrained triaxial tests R1, R2, R3, U1, U2, and U3. There is a reasonably close agreement between the curve and data from four torvane strength tests on undisturbed samples of the UpB till and from 16 shear box tests that are used as an independent check on the best fit relationship. *Source:* Tulaczyk et al. (2000).

The water balance formed another part of the evidence for till deformation at Upstream B. Alley et al. (1986) proposed that if till is not deforming then the ice stream velocity must arise from sliding of ice over a rigid substrate lubricated by water. Water plays a crucial role in the onset of ice streaming and, on a wider scale, the stability of an ice sheet. It can contribute to both till deformation, discussed previously, and basal sliding at the ice-bed interface (Figure 4). The basal sliding process has been examined most extensively on temperate glaciers, which have a thermal regime above the pressure melting point (Bindschadler et al., 2001). The basal thermal regime of ice sheets is important as, it not only affects the relative contribution of internal ice deformation to overall ice stream flow (Hughes, 1977; Tulaczyk et al., 2000), but also the contribution of basal sliding (Weertman, 1957; Payne, 1995). The pressure melting and production of meltwater at the ice stream base is a necessary condition in order to separate the ice from the bed and create fast flow conditions.

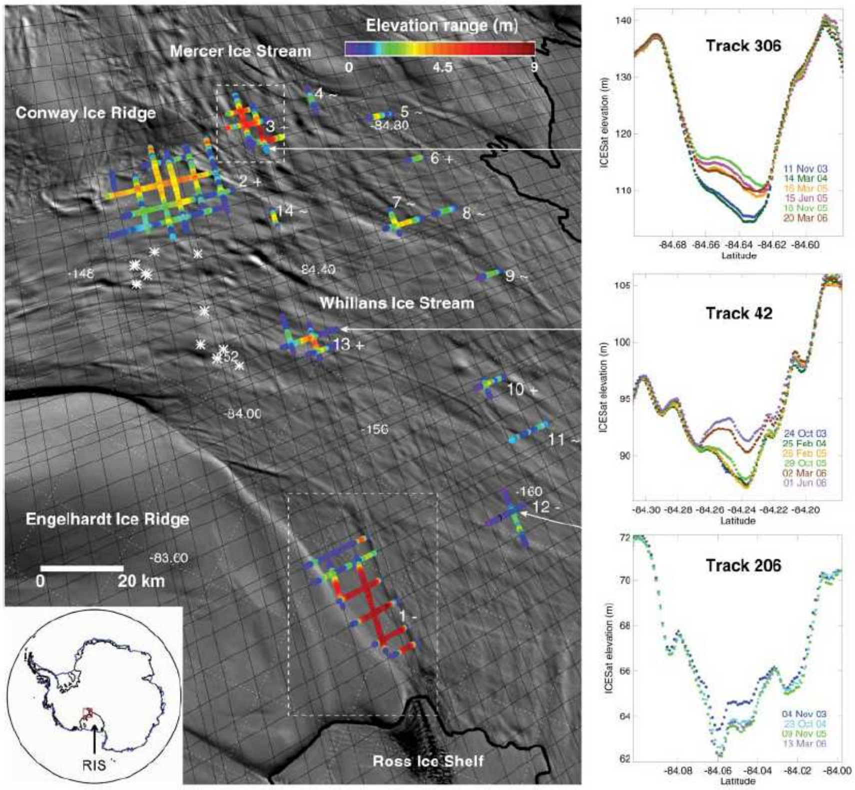

Figure 6: Locations of elevation-change events identified through ICESat repeat-track analysis. Straight black lines show the ICESat reference ground tracks. Coloured track segments represent range in elevation amplitude for each elevation-change event. Events cluster into 14 elevation-change regions, which are either rising (+), falling (-), or oscillating (~). Ice flow is from top left to lower right. Bold black line indicates the break-in-slope associated with the grounding zone of the Ross Ice Shelf. *Source:* Ficker et al. (2007).

Weertman (1969) estimated that fast sliding, which is found in ice streams, requires a water film of thickness $\simeq$ 5 mm or greater. Alley et al. (1986) considered this level of production unlikely due to evidence produced from various studies. Research conducted by Gow et al. (1979) on the catchment area of Ice Stream D showed that net basal freezing rather than melting occurs over some distance up-glacier from the Byrd borehole. This indicates that the basal thermal regime is below the pressure melting point. The catchment of Ice Stream B is quite similar to that of Ice Stream D in surface mass balance, bed topography, surface topography and ice thickness (Drewry, 1983) therefore Alley et al. (1986) expected net basal freezing in parts of the Ice Stream B catchment, however this could vary over both temporal and spatial scales. It appears that ice streams can switch between fast and slow modes of flow through the fluctuation of basal water pressure; however the controls on this are not fully understood (Clarke, 1987; Engelhardt and Kamb, 1998).

Another source of evidence used to discount the production of water at upstream B by Alley et al. (1986) was the research by Budd et al. (1971) who generated a melt rate of 2 mm yr^{-1} using numerical experiments for most of the upstream B catchment. However, this involved increasing the geothermal heat flow by 50% above the value they expected based on the geology of West Antarctica; this value was also confirmed by measurements in the Byrd Station borehole (Rose, 1979). The last form of evidence is in relation to the ≥1 km layer of sedimentary rock which is beneath ice stream B. Alley et al. (1986) state that it may have sufficient permeability to drain a significant amount of water. This suggestion may not be valid as Fricker et al. (2007) has recently discovered that subglacial water is moving in large volumes between lakes, on short time scales and over long distances in the Siple Coast Ice Streams (Joughin et al., 2002; Gray et al., 2005). This was indicated by observed changes in elevation which were interpreted as the surface expression of subglacial water movement (Figure 6). The basal water conduit system appears to contribute to the variation in ice stream flow on both temporal and spatial scales. An example of this is the Kamb Ice Stream where the drastic reduction of flow has been attributed to a decrease in subglacial water supply (Alley et al., 1994).

4.1 Conclusions

Studies on the WAIS-RISS ice streams generally show that fast ice streaming is caused by an efficient basal lubricant and a weak basal till. Because of this weakness of the bed, a significant part of the gravitational driving stress appears to be borne by ice stream margins. Increased knowledge of subglacial water movement is fundamental to understanding Antarctic ice stream dynamics and to predicting ice sheet behaviour and sea-level contribution. Due to the observed complexity of ice streaming, the relative contribution of these two mechanisms to ice flow, over both the temporal and spatial scales for individual ice streams, continues to remain a source for debate.

5.1 References

Alley, R.B., Anandakrishnan, S., Bentley, C.R. and Lord, N. 1994. A water-piracy hypothesis for the stagnation of ice stream C, Antarctica. *Annals of Glaciology* 20, 187-194.

Alley, R.B., Blankenship, D.D., Bentley, C.R. and Rooney, S.T. 1986. Deformation of till beneath ice stream B, West Antarctica. *Nature* 322, 57-59.

Bentley, C.R. 1987. Antarctic ice streams: a review. *Journal of Geophysical Research* 92 (B9), 8843-8858.

Bentley, C.R. 1998. Rapid sea-level rise from a West Antarctic ice-sheet collapse: a short-term perspective. *Journal of Glaciology* 44 (146), 157-163.

Bindschadler, R.A. and Scambos, T.A. 1991. Satellite-image-derived velocity field of an Antarctic ice stream. *Science* 252 (5003), 242-246.

Bindschadler, R., Bamber, J. and Anandakrishnan, S. 2001. Onset of streaming flow in the Siple Coast Region, West Antarctica. The West Antarctic Ice Sheet: Behaviour and Environment, *Antarctic Research Series* 77, 123-136.

Blankenship, D.D., Bentley, C.R., Rooney, S.T. and Alley, R.B. 1986. Seismic measurements reveal a saturated porous layer beneath an active Antarctic ice stream. *Nature* 322, 54-57.

Boulton, G.S. and Paul, M.A. 1976. The influence of genetic processes on some geotechnical properties of glacial tills. *Quarterly Journal of Engineering Geology and Hydrogeology* 9 (3), 159-194.

Boulton, G.S. 1979. Processes of glacier erosion on different substrata. *Journal of Glaciology* 23, 15-38.

Budd, W.F., Jenssen, D. and Radok, U. 1971. Derived physical characteristics of the Antarctic Ice Sheet. *ANARE Interim Reports, Series A (IV) (Glaciology)* 120.

Clarke, G.K.C. 1987. Fast glacier flow: ice streams, surging, and tidewater glaciers. *Journal of Geophysical Research* 92 (B9), 8835-8841.

Drewry, D.J. (ed.) 1983. *Antarctica: Glaciological and Geophysical Folio.* Scott Polar Research Institute, Cambridge.

Engelhardt, H., Humphrey, N., Kamb, B. and Fahnestock, M. 1990. Physical conditions at the base of a fast moving Antarctic ice stream. *Science* 248 (4951), 57-59.

Engelhardt, H. and Kamb, B. 1998. Basal sliding of Ice Stream B, West Antarctica. *Journal of Glaciology* 44 (147), 223-230.

Fahnestock, M. and Bamber, J. 2001. Morphology and surface characteristics of the West Antarctic Ice Sheet. The West Antarctic Ice Sheet: Behaviour and Environment, *Antarctic Research Series* 77, 123-136.

Fricker, H.A., Scambos, T., Bindschadler, R. and Padman, L. 2007. An active subglacial water system in West Antarctica mapped from space. *Science* 315, 1544-1548.

Gow, A.J., Epstein, S. and Sheehy, W.J. 1979. On the origin of stratified debris in ice cores from the bottom of the Antarctic ice sheet. *Journal of Glaciology* 23 (89), 185-192.

Gray, L., Joughin, I., Tulacyzk, S., Spikes, V.B., Bindschadler, R. and Jezek, K. 2005. Evidence for subglacial water transport in the West Antarctic Ice Sheet through three-dimensional satellite radar interferometry. *Geophysical Research Letters* 32, 1-12.

Hughes, T.J. 1977. West Antarctic Ice Streams. *Reviews of Geophysics* 15 (1), 1-46.

Joughin, I. and Tulaczyk, S. 2002. Positive mass balance of the Ross ice streams, West Antarctica. *Science* 295, 476-480.

Joughin, I., Tulaczyk, S., Bindschadler, R.A. and Price, S. F. 2002. Changes in West Antarctic ice stream velocities: Observation and analysis. *Journal of Geophysical Research* 107 (B11), 1-22.

Kamb, B. 2001. Basal zone of the West Antarctic Ice Streams and its role in lubrication of their rapid motion. The West Antarctic Ice Sheet: Behaviour and Environment, *Antarctic Research Series* 77, 157-199.

Oppenheimer, M. 1998. Global warming and the stability of the West Antarctic Ice Sheet. *Nature* 393, 325-332.

Payne, A.J. 1995. Limit cycles in the basal thermal regime of ice sheets. *Journal of Geophysical Research* 100 (B3), 4249-4263.

Payne, A.J. 1999. A thermomechanical model of ice flow in West Antarctica. *Climate Dynamics* 15, 115-125.

Retzlaf, R. and Bentley, C.R. 1993. Timing of stagnation of Ice Streams C, West Antarctica, from short-pulse radar studies of buried surface crevasses. *Journal of Glaciology* 133 (3), 553-561.

Rose, K.E. 1979. Characteristics of ice flow in Marie Byrd Land, Antarctica. *Journal of Glaciology* 24 (90), 63-75.

Shabtaie, S. and Bentley, C. 1987. West Antarctic ice streams draining into the Ross Ice Shelf Configuration and mass balance. *Journal of Geophysical Research* 92 (2), 1311-1336.

Swithinbank, C.W.M. 1954. Ice streams. *Polar Record* 7 (48), 185-186.

Tulaczyk, S., Kamb, B. and Engelhardt, H.F. 2000b. Basal mechanics of Ice Stream B, West Antarctica: 2. Undrained plastic bed model. *Journal of Geophysical Research* 105 (B1), 483-494.

Weertman, J. 1957. On the sliding of glaciers. *Journal of Glaciology* 3 (21), 33-38.

Weertman, J. 1969. Water lubrication mechanism of glacier surges. *Canadian Journal of Earth Sciences* 6, 929-939.

Weertman, J. 1974. Stability of the junction of an ice sheet and an ice shelf. *Journal of Glaciology* 13 (67), 3-11.

Whillans, I.M. and van der Veen, C.J. 1993. New and improved determinations of velocity of Ice Streams B and C, West Antarctica. *Journal of Glaciology* 39 (133), 483-490.

Whillans, I.M. and van der Venn, C.J. 1997. The role of lateral drag in the dynamics of Ice Stream B, Antarctica. *Journal of Glaciology* 43 (144), 231-237.

Note: Image on front page from Scott Polar Research Institute web page www.spri.cam.ac.uk/.../pineislandglacier/ downloaded on 8th December 2008.